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Craton stability and longevity: the roles of composition-dependent rheology and buoyancy

Hongliang Wang¹, Jeroen van Hunen¹, D. Graham Pearson², Mark B. Allen¹

1 Department of Earth Sciences, Durham University, Durham, UK

2 Department of Earth & Atmospheric Sciences, University of Alberta, Edmonton, Canada

Hongliang Wang Email: hongliang.wang@dur.ac.uk

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Abstract:

Survival of thick cratonic roots in a vigorously convecting mantle system for billions of years has long been studied by the geodynamical community. High strength of the cratonic root is generally considered to be the most important factor, but the role of lithospheric mantle depletion and dehydration in this strengthening is still debated. Geodynamical models often argue for a significant strength or buoyancy contrast between cratonic and non-cratonic mantle lithosphere, induced by mantle depletion and dehydration. But recent laboratory experiments argue for only a modest effect of dehydration strengthening. Can we reconcile laboratory experiments and geodynamical models?

We perform and discuss new numerical models to investigate craton stability and longevity with different composition-dependent rheology and buoyancy. Our results show that highly viscous and possibly buoyant cratonic root is essential to stabilise a geometry in which thick cratonic lithosphere and thinner non-cratonic lithosphere coexist for billions of years. Using non-Newtonian rheology, a modest strengthening factor of $\Delta\eta=3$ can protect compositionally buoyant cratonic roots from erosion by mantle convection for over billions of years. A larger strengthening factor ($\Delta\eta=10$) can maintain long term craton stability even with little or no intrinsic buoyancy. Such composition-dependent rheology is comparable to the laboratory experiments. This implies that a strict isopycnic state of cratonic lithosphere may not be necessary for the preservation of a cratonic root, provided a sufficient level of compositional strengthening is present.

1 Introduction

Cratons – the most ancient parts of the continents, are characterized by their old crust and thick lithosphere. Most cratonic lithosphere (e.g. that beneath the Kaapvaal, Siberia and Slave cratons) has a thickness up to 200-250 km (Boyd, 1973; Boyd et al., 1997; Gung et al., 2003; Mather et al., 2011), although some cratons have been subjected to recent thinning events, e.g. the North China Craton (Fan and Menzies, 1992; Gao et al., 2008; Goes et al., 2002). While the oceanic plates recycle into the mantle within about 200 Myrs, cratons are found to be covered by Archean crust. Furthermore, geochemical evidence, predominantly from diamond inclusion dating and Re-Os isotopes, suggests that the age of cratonic lithosphere is broadly similar to its crust, at least to the extent that Archean mantle lithosphere generally underlies Archean crust (Pearson, 1999; Richardson et al., 1984). Hence, it is clear that the present cratonic lithosphere had a thick cratonic keel and survived erosion by mantle convection for billions of years.

While the evidence for the long-term stability of deep cratonic lithosphere is clear, our understanding of all the parameters that contribute to this stability remains relatively poor and conflicting. The longevity and stability of cratons has been attributed to either external factors such as weak mobile belts surrounding cratons, or internal factors such as the intrinsic buoyancy and high strength of the cratonic root (Carlson et al., 2005). Mobile belts surrounding a craton may shield it from plate-tectonic related stresses and erosion, and can increase its stability significantly (Lenardic et al., 2000, 2003; Yoshida, 2010,2012; Yoshida and Santosh, 2011). But for more continuous thermal erosion by the convecting mantle within the asthenosphere, the effects of peripheral mobile belts appear to be limited. This process, known as entrainment of cratonic material (Sleep, 2003) by the lateral and sublithospheric small scale convection, is driven by the lateral and vertical density contrasts surrounding the cratonic root (Doin, et al 1997; Shapiro et al, 1999, Solomatov and Moresi, 2000). Hence, the role of compositionally distinct cratonic root needs to be investigated. Chemical buoyancy promotes resistance to thermal erosion and improves the stability of cratons, but has been suggested to be insufficient to ensure craton longevity by itself (Doin et al., 1997; François et al., 2012; Lenardic, 1999; O'Neill et al., 2008). Thus, “physical protection” from both mobile belts and

chemical buoyancy does not appear to be sufficient to explain why the cratonic root has survived in a convecting mantle for billions of years.

Lenardic et al. (2003) proposed that a thick chemically distinct root with more than twice the thickness of typical oceanic lithosphere is required for long term craton stability even if a very viscous cratonic root (1000 times more viscous) is employed. They further argued that the high brittle yield strength for cratonic lithosphere is a more effective and robust way to provide craton stability over several mantle overturn times. O'Neill et al. (2008) updated this conclusion by demonstrating that long term craton stability can be provided by a compositional viscosity ratio of 50 to 150, with a root /oceanic lithosphere yield strength ratio of 5-30. Based on a scaling research, Sleep (2003) indicates that a viscosity factor of 20 with weakly temperature dependent viscosity is needed for survival of cratonic lithosphere. Furthermore, Beuchert et al. (2010) argued that thermal craton stability in stagnant lid approach for billions years could be provided by sufficiently high temperature-dependent viscosity contrast (10^7 to 10^{10}) even without invoking any compositional difference. While it is well known from petrological studies that the cratonic lithosphere is compositionally distinct from oceanic lithospheric mantle and even post-Archean continental lithospheric mantle (Boyd, 1989; Pearson and Wittig, 2008), the conflicting modelling results show that it remains far from clear to what extent composition-dependent rheology is required. Also, because any mechanism for creating lithosphere creates strong compositional stratification, it is important to explore the role played by a composition-dependent rheology and buoyancy.

The need to further explore the effect of composition-dependent rheology on lithosphere stability is also driven by results from laboratory rheological experiments. Volatiles, especially water are supposed to be among the dominant influences on mantle rheology (Karato, 2010; Keefner et al., 2011; Mei, 2000a, 2000b). The dehydration of cratonic roots due to their exceptional melt depletion has long been proposed to contribute to the high strength of cratonic lithosphere (Carlson et al., 2005; Hirth et al., 2000; Peslier et al., 2010). But the effects of water on mantle rheology in laboratory experiments vary considerably between different experiments (Fei et al., 2013; Hirth and Kohlstedt, 1996; Karato, 2010; Mei, 2000a, 2000b). Hirth and Kohlstedt (1996) found the viscosity of water-

saturated olivine aggregates is reduced by a factor of 100~180 at a confining pressure of 300 MPa (where the solubility of water is ~15.3wt ppm or $\sim 250\text{H}/10^6\text{Si}$). Karato (2010) proposed viscosity changes up to 4 orders of magnitude for a constant stress at depths of 200-400 km due to the influence of water. In contrast, a recent study of the silicon self-diffusion coefficient in forsterite suggests that the effect of water on upper mantle rheology is only up to one order of magnitude (Fei et al., 2013).

The role of different deformation mechanisms in the upper mantle might also need further investigation. Most of the previous numerical research on craton stability used a Newtonian rheology for mantle deformation. François et al. (2012) used non-Newtonian rheology but without any compositional dependency. Mantle rheology laboratory studies suggest that the dominant creep mechanism in upper mantle is dislocation creep, even though diffusion creep (grain-size sensitive) may operate in some conditions (Karato, 2010). This point has also been supported by geodynamical modelling and measurements of seismic anisotropy of upper mantle (Mainprice et al., 2005; van Hunen et al., 2005; van Hunen and Čadež, 2009). Diffusion creep, usually taken as Newtonian rheology in numerical modelling, could turn into non-Newtonian rheology if the grain-size dependence is not ignored and influence the mantle flow pattern substantially (Rozel et al., 2012). Although Newtonian rheology is proved to be able to mimic non-Newtonian rheology with an equivalent activation energy (Christensen, 1984), it may fail to capture some important characteristics of mantle convection (e.g. Asaadi et al., 2011).

Given the significant variation in results from both geodynamical modelling and laboratory rheological experiments, further investigation in the effects of compositional rheology and buoyancy due to mantle dehydration and depletion on craton stability is warranted. Here, we present numerical modelling results on the effects of compositional rheology and buoyancy on the stability of cratonic roots. We aim to determine how viscous or buoyant cratonic roots need to be to ensure craton stability and longevity over Gyr timescales using both non-Newtonian and Newtonian rheology.

2 Model description

2.1 Governing equations

A Cartesian version of the finite element code Citcom is used in our numerical simulation (Moresi and Solomatov, 1995; van Hunen and Allen, 2011; Zhong et al., 2000), which solves the incompressible flow using the Boussinesq approximations. The non-dimensional governing equations of conservation of mass, momentum, energy are:

$$\nabla \cdot \mathbf{u} = 0 \quad (1)$$

$$-\nabla P + \nabla \cdot (\eta (\nabla \mathbf{u} + \nabla^T \mathbf{u})) + (RaT - Rb_i C_i) \mathbf{e}_z = 0 \quad (2)$$

$$\frac{\partial T}{\partial t} + (\mathbf{u} \cdot \nabla) T = \nabla^2 T + Q_0$$

A standard non-dimensionalisation is applied, with $x = x' h, t = t' h^2 / \kappa, \eta = \eta' \eta_0$. The primes of the non-dimensional parameters are dropped in the equations above for clarity. Their dimensional counterparts are defined in Table 1. Different chemical fields are used to track different material (C_1 for upper crust and lower crust; C_2 for depleted mantle and fertile mantle), which are advected with the flow through the equation:

$$\frac{\partial C_i}{\partial t} + (\mathbf{u} \cdot \nabla) C_i = 0$$

(4)

where C_i is calculated from the distribution of tracers. Eq. (4) is solved by a particle-tracking technique (Ballmer et al., 2007), in which the tracers, that carry the information about density and rheology, are advected with the velocity field from Eq. (1) and (2). This method has been benchmarked against van Keken et al. (1997) and Schmeling et al. (2008). One tracer function is used to track crustal material (including upper and lower crust), while another tracks different depletion of mantle material (fertile mantle, highly depleted, and less depleted cratonic root material) (Figure 1).

The thermal Rayleigh number Ra and compositional Rayleigh number Rb are defined as:

$$Ra = \frac{\alpha \rho_0 g \Delta T h^3}{\kappa \eta_0} \quad (5)$$

$$Rb_i = \frac{\delta \rho_i g h^3}{\kappa \eta_0} \quad (6)$$

The description of parameters in Eqs. (5)-(6) is listed in Table 1. The Rayleigh numbers given in the Table are calculated by using relatively high reference viscosity η_0 . The viscosity of asthenosphere is, however, significantly lower (e.g. Figure 2).

2.2 Model setup

Our models extend to 660 km in depth and 2640 km in horizontal width, with a 660 km-wide craton positioned in the middle of the model domain (Figure 1). In order to deal with locally sharp and large viscosity contrast, a resolution of 192-by-96 finite elements is used, with mesh refinement in the cratonic root area. This provides the vertical resolution is 4.7 km around the cratonic root area and 8-9 km in the circum-cratonic regions. 640000 tracers were used in the model, which provides an average tracer density of about 35 tracers per finite element. The velocity boundary conditions used in the models are free slip at the surface and side boundaries, and no-slip at the bottom. Half space cooling models are used to define the initial thermal structure of cratonic lithosphere (400 Myrs in most cases) and normal lithosphere (150 Myrs). Temperature is fixed at the surface ($T=0^\circ\text{C}$) and bottom ($T_b=1485^\circ\text{C}$ in most cases), while side boundaries are insulating. This setup leads to a mantle potential temperature around 1350°C in asthenosphere after the thermal boundary layers form at bottom in our models. In order to monitor the thickness changes of the lithosphere, we define a thermal lithosphere asthenosphere boundary (LAB) by the isotherm of $T=1215^\circ\text{C}$ and calculate the average thicknesses of cratonic and non-cratonic lithosphere through time. The selection of rheological parameters in most models (all except N11 and N12) is based on the assumption that non-cratonic lithosphere should have a thickness of about 140 km. This value is a reasonable average for

Phanerozoic and Neoproterozoic lithosphere, i.e. lithosphere younger than ~1 Ga (Artemieva, 2011). These rheological parameters also result in a reasonable viscosity at asthenosphere between 10^{18} Pa·s and 10^{19} Pa·s that is constant with the value suggested by Solomatov and Moresi, 2000.

A two-layer cratonic keel model (modified from Afonso et al. (2008)'s three-layer model) was used: a highly depleted layer (with $C=1$) from 36 km (Moho) to 144 km depth and a less depleted layer ($C=0.6$) from 144 km to 234 km depth. We assume the density reduction to be proportional to C , with the most depleted lithosphere 2.1% (70 kg/m^3) less dense than fertile lithosphere, which is within the range of density reduction (1.5% ~2.5%) due to a bulk Mg number change from fertile mantle (e.g. $\text{Mg\#}=88$) to depleted cratonic root ($\text{Mg\#}=92 - 94$) (Pearson and Wittig, 2008). The density of crust and the depleted cratonic root is described by:

$$\rho_c = \rho_m - C_i \Delta \rho_i \quad (7)$$

All models have a buoyant crust composed of 18 km upper crust and 18 km lower crust, which are 600 kg/m^3 and 400 kg/m^3 less dense than the normal mantle, respectively. However, we found that, the effect of this crust density contrast on the dynamics of the convection system is minor. We ignore the difference in radioactive heating between crust and mantle and their decay over the long-term history of the Earth, but use a uniform, equivalent value of $Q_0 = 0.04 \text{ } \mu\text{W/m}^3$, which is roughly twice the value of modern, undepleted mantle as an average for the past 2 Ga. This simplified heat production model allows us to investigate the statistically steady-state situation, and therefore gain a more fundamental understanding of the dominant processes.

2.3 Rheology

As this study focuses on the role played by the rheology of cratonic root, we ignore any rheological differences made by the crust to simplify the model setup. The effective viscosity in our models is described by:

$$\eta_m = A \cdot \frac{1-n}{\dot{\epsilon}^n} \exp\left(\frac{E + \rho g z V}{nRT}\right) \quad (8)$$

with $n=3.5$ for non-Newtonian rheology (i.e. dislocation creep) and $n=1$ for Newtonian rheology (i.e. diffusion creep). Our non-Newtonian rheology has been benchmarked against results of Christensen (1984). The descriptions and default values of the parameters in Eq. (8) are listed in Table 1.

In order to study the effects of composition-dependent rheology of cratonic root, a strengthening factor $\Delta\eta$ is used. As a result, the effective viscosity of cratonic root is dependent on the depletion, and given by:

$$\eta_c = \eta_m \times \Delta\eta^{\min(1, \frac{C_2}{C_\eta})} \quad (9)$$

Combined with Eq. (8), Eq. (9) defines a composition-dependent rheology in which the effective viscosity of cratonic root is $\Delta\eta$ times of normal mantle at constant strain rate. This is equivalent to a viscosity contrast $\Delta\eta^n$ if the constant-stress definition of effective viscosity is used for non-Newtonian rheology. A maximum model viscosity cut-off was set up to 10^{25} Pa s in most models, but tests with 10^{26} Pa s and 10^{27} Pa s were done to assure that higher viscosity cut-off would not affect the results.

3 Results

To investigate cratonic stability under a range of circumstances, numerical model calculations were performed with different rheology and buoyancy settings (Table 2). We define a successful model for craton preservation using the following two criteria: 1) the cratonic root survives erosion via mantle convection and preserves a thickness in excess of 200 km for 2 Gyrs; 2) over the same period, non-cratonic lithosphere maintains a reasonable thickness of $\sim 140 \pm 10$ km. To illustrate the dominant geodynamical features in those models, we start with

the description of a reference model.

3.1 Reference model

Our reference model uses the two-layered cratonic root and non-Newtonian rheology with a compositional rheological factor $\Delta\eta=10$ (Table 1, 2). Figure 2 shows the evolutions of temperature, cratonic root shape and viscosity for the duration of 2 Gyrs. The main part of the cratonic root is stable, with only limited erosion at its edges. There is a viscosity contrast of about 3 to 4 orders magnitude between the cratonic root and asthenosphere (Figures 2B, 2D, 2F). Hence, due to the combined non-Newtonian and temperature dependent effect, the compositional rheological factor $\Delta\eta=10$ results in sufficient viscosity at the cratonic root to resist significant erosion from the underlying convecting mantle.

The thickness changes of cratonic lithosphere and non-cratonic lithosphere are monitored through time by using the $T=1215^{\circ}\text{C}$ isotherm as the LAB. The results are plotted as the red lines in figure 3A, with thick and thin lines represent cratonic and non-cratonic lithosphere respectively. In this model, the thicknesses of cratonic and non-cratonic lithosphere are maintained as 210 km and 140 km respectively.

3.2 Compositional rheology

Figure 3A shows four numerical experiments performed with different strengthening factors $\Delta\eta$ in terms of thickness evolution of cratonic and non-cratonic lithosphere, including reference model N3. All of the other three models have the same initial buoyant cratonic root setting as in the reference model. Without any compositional rheology (model N1, $\Delta\eta=1$), the thickness of cratonic lithosphere decreases slowly with time (Figure 3A), and cratonic root almost disappears over 1.2 Gyrs (Figure 4A). This means that just starting with a buoyant and cold cratonic keel (and therefore stronger than the surroundings) is not sufficient to preserve a

craton from long term heating and erosion. An even higher strengthening factor (model N4, $\Delta\eta=30$) restricts the erosion of cratonic lithosphere only marginally more than the default $\Delta\eta=10$ (Figure 3A). Comparably, a smaller strengthening factor (N2, $\Delta\eta=3$) may also maintain the thick cratonic root for billions of years (Figure 3A), but the lowermost cratonic root is stretched significantly (Figure 4B).

From Figure 3A, an estimate of the thinning rate of cratonic lithosphere is calculated through polynomial fitting. Models with composition-dependent rheology show a substantial reduced thinning rate compared to model N1. The thinning rate in the reference model N3 ($\Delta\eta=10$) decreases from $\sim 30\text{km/Gyrs}$ in the beginning to $\sim 2\text{km/Gyrs}$ at 2 Gyr, which makes a well-preserved craton. The thinning rate in model N1 ($\Delta\eta=1$) also decreases with time, but it start at $\sim 80\text{km/Gyrs}$ and maintain a rate of $\sim 20\text{km/Gyrs}$ at 1.5Gyr, which precludes the longevity of thick cratonic lithosphere. This shows that a strong cratonic root is essential to resist the long-term erosion from the convecting mantle

Another 4 similar experiments were performed but without compositional buoyancy in the cratonic roots (Figures 3B, 4C, 4D). Despite the absence of compositional buoyancy, the cratonic lithosphere is still well preserved with $\Delta\eta=10$ in model N7, and keeps a thickness well beyond 200 km throughout the entire 2 Gyrs model calculation (Figures 3B, 4D). Again, the difference between $\Delta\eta=30$ (model N8) and $\Delta\eta=10$ (model N7) is statistically insignificant. Thus, no higher strengthening factor than $\Delta\eta=10$ is required even if the cratonic root is not compositionally buoyant.

3.3 Compositional buoyancy

The effects of compositional buoyancy are investigated by the comparison between models with and without buoyant cratonic roots (model N1-N4 in figure 3A versus N5-N8 in figure 3B). Model N5 shows that the cratonic lithosphere would be eroded away rapidly within 400

Myrs (Figure 3B) if the compositional buoyancy is removed from model N1. Model N6 demonstrates that the smaller strengthening factor ($\Delta\eta=3$) cannot prevent substantial entrainment of cratonic material if the cratonic root is not compositionally buoyant (Figures 3B, 4C). This illustrates that compositional buoyancy could cancel the negative thermal buoyancy of cratonic root and thus keep cratonic material stay underneath the area surrounding the craton (Figures 4A, 4B). The changes in the geometry of cratonic root in model N7 (Figure 4D) also shows the compositional buoyancy has some effects on the long-term deformation of cratonic roots. However, due to the lack of sufficient negative thermal buoyancy to counteract the compositional buoyancy, the lowermost cratonic root may suffer from an extensional gravity instability that stretches the root and thus smoothes the boundary between cratonic and non-cratonic lithosphere as Figure 4B shows. Hence, the compositional buoyancy of the cratonic root plays a dual role: while compositional buoyancy can help resist the entrainment of cratonic material into asthenosphere significantly and contribute to the longevity of thick cratons, it may also reduce the craton thickness via extensional gravity instability in the lowermost root.

Despite the significant effect of compositional buoyancy in some of the models discussed here, it is unlikely to play a dominant role in the survival of cratons for billions of years. Compositional buoyancy has important influence in models with less strong cratonic roots (model N1, N2), but the geometry of cratonic root is not well preserved in these models. The models with $\Delta\eta=10$ or more have the capacity to maintain long-living cratons, but in those models, the role of compositional buoyancy is rather modest.

3.4 Mantle viscosity

Long term craton stability may not only be affected by the strength of the cratonic root itself,

but also by the viscous properties of the surrounding mantle and lithosphere (Doin et al., 1997; Beuchert et al., 2010). In this section, models are designed to investigate the effects of pressure-dependent viscosity and the maximum lithospheric viscosity cut-off. Additionally, experiments with less viscous mantle and a warmer mantle are also performed in order to explore the parameter sensitivity.

Relative to reference model N3, we first increase the activation volume from 5 cm³/mol to 10 and 15 cm³/mol in model N9 and N10. In order to keep non-cratonic lithosphere at around 140 km, the basal temperature and rheological parameter A were adjusted accordingly. Higher activation volume V induces stronger variations of lithosphere thickness, but a thickness difference of ~70km between cratonic and non-cratonic lithosphere is still maintained (Figure 5A). Higher V increases the viscosity of the deeper mantle, which results in a thicker basal thermal boundary layer and larger, less frequent, but more regular upwelling thermal structures forming under the cratons, which, in turn, tend to erode the cratonic root slightly (Figures 2, 6A, 6C). This induces a periodic pattern of cratonic root thickness variations in model N10 (Figure 5A). The shape of the cratonic roots deforms due to these upwelling thermal structures shown in Figure 6A and 6C, but they do not thin the cratonic lithosphere significantly. Hence, a strengthening factor of 10 with non-Newtonian rheology could still ensure long term craton stability even with stronger pressure dependence of the viscosity.

To investigate the effects of viscosity contrast, model N11 and N12 were performed with a maximum viscosity changed from 10²⁵ Pa·s to 10²⁴ Pa·s and 10²⁶ Pa·s, respectively (Figure 5B). While the non-cratonic lithosphere keeps the same thickness as in the reference model, a higher maximum viscosity makes only marginal difference to cratonic lithosphere, whereas a lower maximum viscosity leads to a significant thinning of cratonic thickness (Figure 5B). This confirms that the default viscosity contrast in our model is sufficiently high

for long term craton stability, that more elevated values will not induce a substantial difference, and that a significantly lower value would not be sufficient to maintain craton stability for billions of years.

Model N13 and N14 in Figure 5C illustrate the effects of a more vigorous and warmer mantle, respectively, conditions that may be applicable to the Archean Earth (Herzberg et al., 2010). Both cratonic and non-cratonic lithosphere are thinner than in the reference model N3, but the thinning of cratonic lithosphere (~10-15 km) is significantly smaller than that of non-cratonic lithosphere (~30-40 km) (Figure 5C, 6B, 6D). This shows that a difference in thickness between two types of lithosphere is expected to be maintained (or even enhanced) in a more vigorous mantle.

3.5 Newtonian rheology

All previous models were calculated using a non-Newtonian rheology with typical dislocation creep parameters. To compare with previous studies and to put the non-Newtonian results in context, we performed several Newtonian calculations. The activation energy is chosen within the range from laboratory experiment, and the rheological pre-factor A is adjusted to assure that non-cratonic lithosphere maintains a thickness of ~140 km (Table 2). Figure 7 shows the lithosphere thickness evolution in Newtonian models (L1 and L2), together with the reference model N3 for comparison.

A modest activation volume of $V=2.5 \text{ cm}^3/\text{mol}$ is used in model L1 and L2 to introduce a pressure-dependent rheology and a slightly more viscous transition zone. In Newtonian rheology, $\Delta\eta=10$ cannot maintain a cratonic root whereas $\Delta\eta=100$ enables the cratonic root to maintain a thickness of ~190 km for 2 Gyrs (Figure 7). This required viscosity contrast for Newtonian rheology agrees with the findings by O'Neill et al. (2008),

who found that a viscosity ratio of 50~150 is required between the cratonic root and asthenosphere when Newtonian rheology is applied. However, our reference model N3 and even model N2 ($\Delta\eta=10$ and $\Delta\eta=3$ with non-Newtonian rheology) still preserve cratonic lithosphere better than model L2 ($\Delta\eta=100$ with Newtonian rheology) according to the thickness evolutions of cratonic lithosphere (Figures 3A, 7).

4 Discussion

4.1 Comparison with previous work

Our geodynamical models illustrate the critical role of compositionally dependent rheology and buoyancy in the long-term stabilisation of thick Archean lithosphere. A strengthening factor of $\Delta\eta=10$ in non-Newtonian rheology can protect the cratonic keel in a stagnant lid situation for billions of years even with little or no compositional buoyancy, while a smaller strengthening factor $\Delta\eta=3$ could also maintain thick but neutrally buoyant cratonic root (isopycnic status). In comparison with laboratory studies on mantle rheology, strengthening factor of $\Delta\eta$ in Eq. (9) would translate into $\Delta\eta^n$ for the constant-stress definition used in the laboratory studies. Therefore, $\Delta\eta=10$ falls into the range of experimental results for dehydration strengthening for dislocation creep between ~140 (Hirth and Kohlstedt, 1996) and 10000 (Karato, 2010). Fei et al. (2013) argue that the effect of water on viscosity of upper mantle is only up to one order of magnitude. Such low values may apply better to our findings that $\Delta\eta=3$ with non-Newtonian rheology might be sufficient to support longevity, and could indicate that buoyancy plays a more significant role.

The role played by composition-dependent rheology has been previously questioned by Beuchert et al. (2010), with the argument that published numerical models have insufficient temperature dependent viscosity contrast, which could drive an artificial requirement for composition dependent rheology. Our research confirms that the viscosity contrast of 10^5 , normally used in many geodynamic modelling, is not suitable for modelling long term craton stability (Figure 5B). However, our models show that composition-dependent rheology is still required to resist the long term erosion

and maintain the different thicknesses between cratonic and non-cratonic lithosphere, and such behaviour is almost inevitable from the observed compositional variation of many cratonic lithospheres (e.g., Pearson & Wittig, 2014). The erosion process in our model, previously described as entrainment of cratonic material, requires high viscosity rather than buoyancy alone for craton preservation as emphasized by Sleep (2003). Using a scaling law relationship, Sleep (2003) concluded that weakly temperature dependent viscosity require lower strengthening factor than strong temperature-dependent viscosity. With non-Newtonian rheology, the temperature dependence of viscosity is reduced by a factor of “n” (3.5) and also by the strain rate dependence compared to Newtonian rheology, which plausibly explain why non-Newtonian rheology require a much smaller $\Delta\eta$ than Newtonian rheology in our models.

4.2 Cratons in a thermally evolving earth

Our models are designed largely from consideration of the current mantle status and result in a strengthening factor of 10 or less is required for long term craton stability. It is widely accepted that the mantle in the Archean was much hotter and thus more vigorous than the current mantle (e.g., Herzberg et al., 2010). A hotter early mantle means that cratons possibly experienced stronger convection and erosion when formed in the Archean and so we need to consider this potential effect. As our models show, the thickness difference of more than 70 km between cratonic and non-cratonic lithosphere could persist in a less viscous upper mantle (Figures 5C, 6B, 6D), even though such a hotter mantle leads to a slightly reduced thickness of both cratonic (10~15 km and non-cratonic lithosphere (30~40 km). Hence, the $\Delta\eta=10$ in our model is able to maintain the thickness difference between cratonic and non-cratonic lithosphere in a hotter and vigorous mantle, thought to be more akin to Archean conditions. More realistic convection parameters of the present-day Earth with plate tectonics and perhaps mantle plumes, might require additional conditions for the survival of cratons, such as weak mobile belts or continental margins (Lenardic et al., 2000, 2003; Yoshida, 2010, 2012). However, this is beyond the scope of this paper.

The compositional buoyancy of cratonic roots could delay the erosion from small-scale convection significantly (Figures 3A and 3B) and has an influence on the long term deformation and shape of cratonic roots (Figures 2 and 4D), but it is not really necessary for craton preservation. The intrinsically more viscous cratonic root would allow modest negative buoyancy of the cratonic root according to our numerical experiments. Modest negative, or variable buoyancy in cratonic roots is consistent with modelling of temperature induced gravity anomalies beneath cratons (Kaban et al., 2003) that indicates that density variations due to temperature are only partly compensated for by density variations due to composition. Such observations along with petrological studies of cratonic mantle xenoliths that show variations in density with depth/equilibration temperature deviating significantly from the expected variation of density with depth for an isopycnic state (Boyd et al., 1997; Lee, 2006; Schutt and Leshner, 2010) indicate that the “isopycnic” hypothesis of Jordan (1978) is not a necessary condition for craton survival. Furthermore, a 500~1500m topographical depression of cratons compared to previously predictions from pure crustal isostasy, has been suggested by crustal isostatic research (Mooney and Vidale, 2003). This, combined with the fact that a number of cratons with thick lithospheric roots such as the Kaapvaal and Siberian cratons are covered to some degree with Phanerozoic oceanic sediments, indicates that craton freeboard might have varied through time. Our modelling has the important implication that the common petrological processes that lead to an increase in the density of the cratonic root after their formation, such as refertilization of peridotites by silicate metasomatism (Simon et al., 2003; Schutt and Leshner, 2010) do not necessarily jeopardize the stability of cratons.

4.3 Implications for thinning of cratonic lithosphere

The process of thinning and removing cratonic lithosphere has generated great attention recently since some cratons have been found to have lost part of, or most of their lithospheric roots (Fan and Menzies, 1992; Foley, 2008; Kusky et al., 2007; Lee et al., 2011). Our models illustrate that the cratonic root should be more viscous and thus very robust against long term erosion by small-scale convection, and explain craton stability and

longevity in a stagnant lid regime. Geodynamical modelling of the process of significant removal of cratonic lithosphere should take this increased viscosity of cratonic root into account. As the cratonic root is more viscous than normal mantle under the same conditions, the heating (from mantle plumes or any other thermal events) would likely have a bigger impact on non-cratonic lithosphere than on cratonic roots. The greater susceptibility of less depleted, non-cratonic mantle to thermal erosion is in line with the observed thickness reduction experienced by circum-cratonic lithosphere in southern Africa (Bell, 2003; Mather et al., 2011). Given the robustness of cratonic roots, a special weakening mechanism is required to destabilize them.

Compared with previous studies, this study shows that non-Newtonian rheology (dislocation creep regime) with compositional strengthening due to mantle depletion and dehydration that is in agreement with laboratory experimental studies is able to ensure long term craton stability. This point, on the other hand, also indicates the cratons could be very susceptible to stress change. Furthermore, recent laboratory experiments show that the interaction between melt segregation and stress would weaken mantle material more than a homogeneous melt distribution (Holtzman et al., 2012). This effect would enhance the stress and strain rate localization during any thermal-tectonic events and make significant removal of cratonic lithosphere possible. Water, despite its debated direct effects on mantle rheology, could also play an important role as it could induce hydrous redox melting by lowering the solidus of mantle rocks and thus weaken cratonic root in that way (Foley, 2008, 2011; Green et al., 2010). These factors, combined with stress effects, need to be investigated geodynamically for their potential to destroy cratonic roots.

5. Conclusion

We have studied the role played by composition-dependent rheology and buoyancy on craton stability and longevity in non-Newtonian rheology. Based on our models, the following points are concluded.

1. Composition-dependent rheology is necessary in order to maintain a different thickness between cratonic and normal lithosphere for over billions of years.

2. With non-Newtonian rheology, a strengthening factor of 10 could preserve a cratonic root from the erosion by small scale convection over 2 Gyrs, no matter whether the cratonic root is compositionally buoyant or not. A smaller strengthening factor ($\Delta\eta=3$) can also protect the a cratonic root from eroding away for billions of years but a neutrally buoyant root is also required in this scenario.

3. The buoyancy of a cratonic root does have some effects on long term cratonic stability, but its effects may be secondary compared to composition-dependent rheology. Positive or even modest negative buoyancy of a cratonic root is permitted, provided that sufficient composition-dependent rheology is present. This conclusion is in agreement with the relatively subtle petrological controls on the buoyancy of mantle rocks and the geological evidence that the freeboard of cratons may have varied considerably through time, with several cratons having been at or below sea level for distinct periods.

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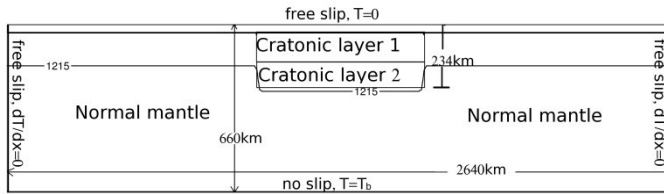


Figure 1. Model setup, including initial material setup, mechanical and thermal boundary conditions. The model extends to 660 km depth and has a 1:4 aspect ratio. The material setup includes: a 36 km crust on the surface, two cratonic layers in the middle of the model domain (extends to 234 km depth); normal mantle elsewhere. The first and second cratonic layers are 70 kg/m^3 and 42 kg/m^3 less dense than the normal mantle, respectively. Half space cooling models are used to define the initial thermal structure of cratonic lithosphere (400 Myrs unless mentioned otherwise) and normal lithosphere (150 Myrs). The curve in the domain indicate the isotherm of and 1215°C .

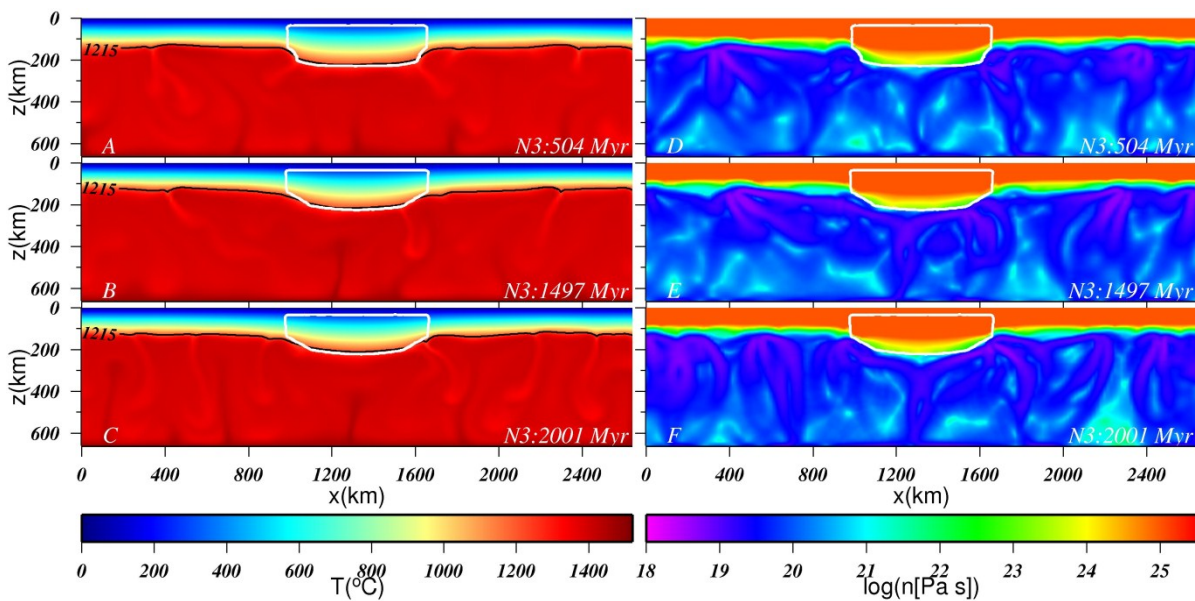


Figure 2) Thermal (a,b,c) and rheological (d,e,f) evolution of the reference model N1 for a model time of 0.5 Gyrs after the start of the model calculation (a,d), 1.5 Gyrs (b,e), and 2.0 Gyrs (c,f). The white contour outlines the compositionally different cratonic root. The isotherm of $T=1215^\circ \text{C}$ is plotted in the temperature image to represent the lithosphere asthenosphere boundary (LAB). The reference model shows cratonic root has been well preserved through about 2 Gyrs.

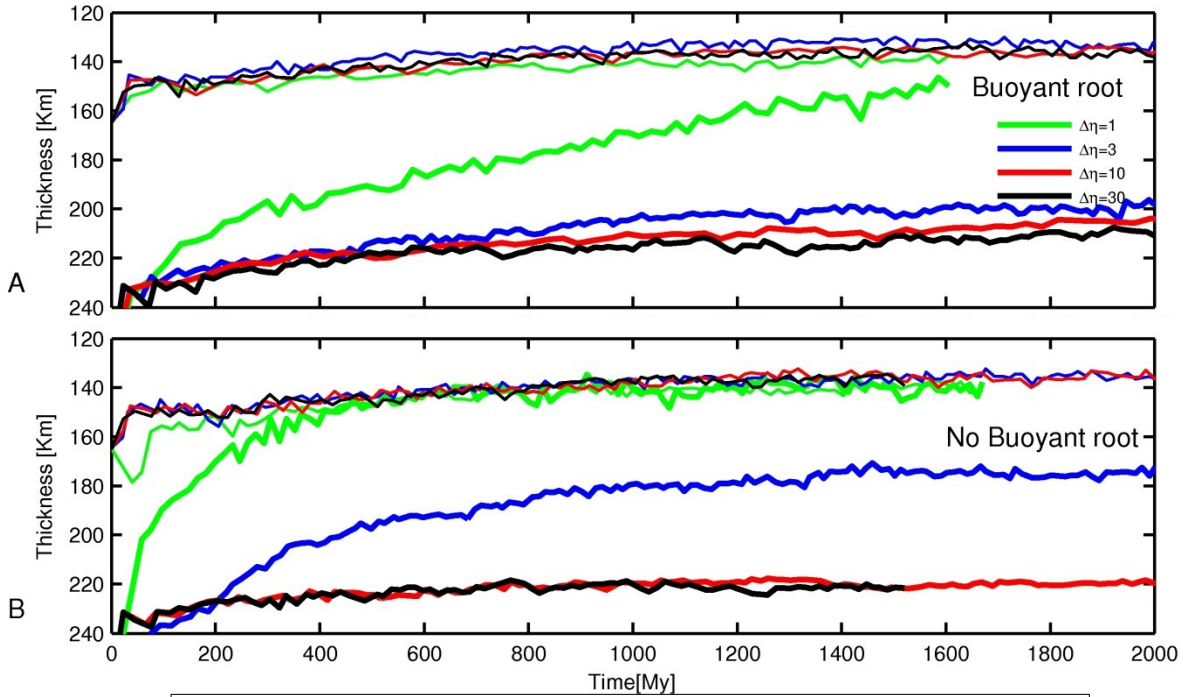


Figure 3 Thickness evolutions of cratonic (thick lines) and non-cratonic lithosphere (thin lines) for models with non-Newtonian rheology. The red lines in 3A represent the reference model N3, while green, blue and black lines represent models with strengthening factor of 1 (model N1), 3 (model N2) and 30 (Model N4). Figure 3B shows models without extra buoyancy of the cratonic root (model N5, N6, N7, N8), compared with models in 3A.

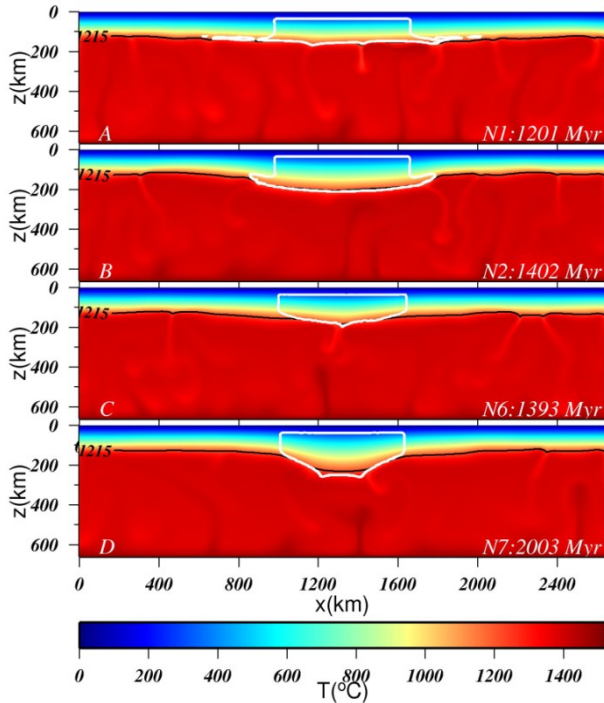


Figure 4. The temperature field and cratonic root (white box) of model N1 at 1.2 Gyr (A), N2 at 1.4 Gyr (B), N6 at 1.4 Gyr (C) and N7 at 2 Gyr (D). Compositional buoyancy of cratonic root delays the erosion process in model N1 (A) compared with model N5. Model N7 has a different shape of cratonic root with the reference model N3, which indicates that compositional buoyancy of cratonic root might have an important effect on the long term deformation of cratonic lithosphere.

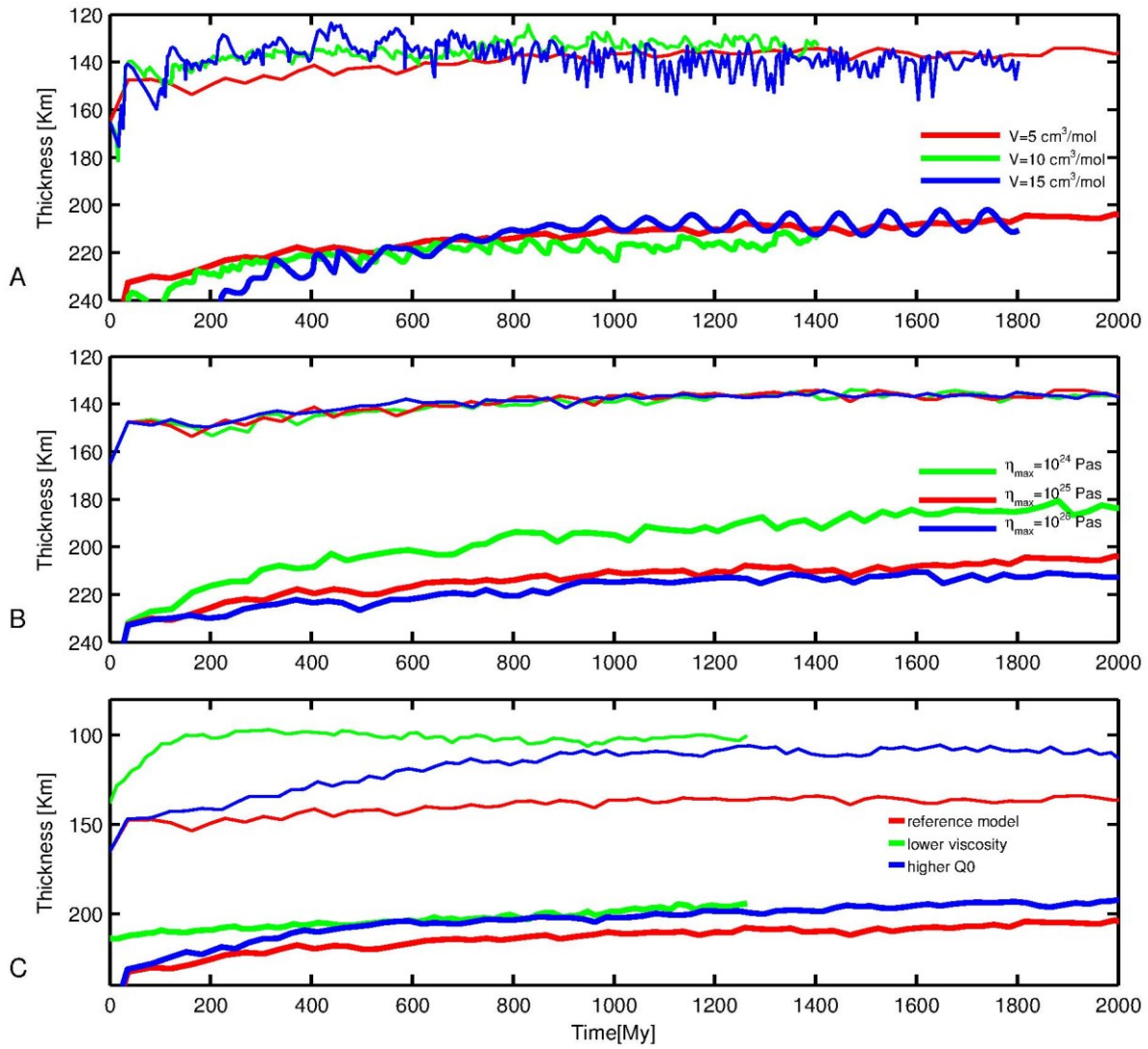


Figure 5 Thickness evolutions of cratonic and non-cratonic lithosphere through 2 Gyrs in different models, compared with the reference model N3. (A): Different activation volumes (models N9, N10). (B): Different maximum viscosity cutoff (models N11, N12). (C): More vigorous mantle by reducing the background viscosity with higher A (model N13) or using higher radioactive heating Q_0 (model N14).

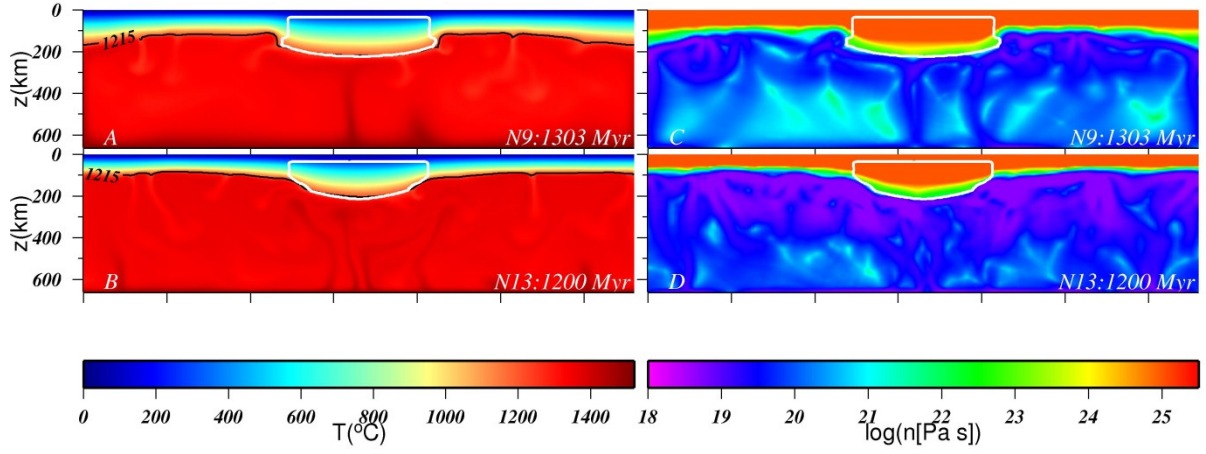


Figure 6. Temperature (A,C) and viscosity (B,D) distribution of model N9 ($V=10 \text{ cm}^3/\text{mol}$) at 1.4 Gyr and model N13 at 1.2 Gyr. Model N9 shows the effects of stronger pressure-dependent rheology, which strengthens the upwelling flow under cratons. Model N13 shows the effects of a more vigorous mantle, in which the thickness of non-cratonic lithosphere decreases much more than cratonic lithosphere.

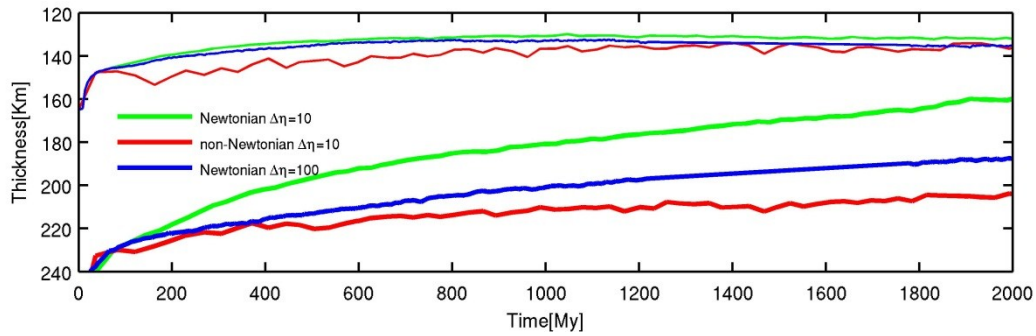


Figure 7. Thickness evolutions of cratonic and normal lithosphere in Newtonian rheology models (L1, L2), compared with reference model N3.

